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# Modelling monthly runoff generation processes following land use changes: groundwater–surface runoff interactions

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## Abstract

A conceptual water balance model is presented to represent changes in monthly water balance following land use changes. Monthly rainfall–runoff, groundwater and soil moisture data from four experimental catchments in Western Australia have been analysed. Two of these catchments, ‘Ernies’ (control, fully forested) and ‘Lemon’ (54% cleared) are in a zone of mean annual rainfall of 725 mm, while ‘Salmon’ (control, fully forested) and ‘Wights’ (100% cleared) are in a zone with mean annual rainfall of 1125 mm. At the Salmon forested control catchment, streamflow comprises surface runoff, base flow and interflow components. In the Wights catchment, cleared of native forest for pasture development, all three components increased, groundwater levels rose significantly and stream zone saturated area increased from 1% to 15% of the catchment area. It took seven years after clearing for the rainfall–runoff generation process to stabilise in 1984. At the Ernies forested control catchment, the permanent groundwater system is 20 m below the stream bed and so does not contribute to streamflow. Following partial clearing of forest in the Lemon catchment, groundwater rose steadily and reached the stream bed by 1987. The streamflow increased in two phases: (i) immediately after clearing due to reduced evapotranspiration, and (ii) through an increase in the groundwater-induced stream zone saturated area after 1987. After analysing all the data available, a conceptual monthly model was created, comprising four inter-connecting stores: (i) an upper zone unsaturated store, (ii) a transient stream zone store, (iii) a lower zone unsaturated store and (iv) a saturated groundwater store. Data such as rooting depth, Leaf Area Index, soil porosity, profile thickness, depth to groundwater, stream length and surface slope were incorporated into the model as *a priori* defined attributes. The catchment average values for different stores were determined through matching observed and predicted monthly hydrographs. The observed and predicted monthly runoff for all catchments matched well with coefficients of determination ( $R^2$ ) ranging from 0.68 to 0.87. Predictions were relatively poor for: (i) the Ernies catchment (lowest rainfall, forested), and (ii) months with very high flows. Overall, the predicted mean annual streamflow was within  $\pm 8\%$  of the observed values.

**Keywords:** monthly streamflow, land use change, conceptual model, data-based approach, groundwater

## Introduction

Climate, soil and vegetation play important roles in maintaining the water balance of catchments (Eagleson, 1978). Changes in even one of these factors will affect the hydrological response. Milly (1994a) showed that the regional annual water balance is primarily a function of climate and to a lesser extent of the soil storage capacity of the landscape. Milly (1994b) later extended this analysis of the annual water balance, presenting seven dimensionless ratios that encapsulate the combined effects of climate, soil and vegetation.

The approach of Milly (1994a,b) has previously been referred to as the ‘*downward approach*’ by Klemes (1983). It differs philosophically from the *upward* or *bottom-up* approach that attempts to combine laboratory-scale understanding of the hydrological processes into mathematical models which are capable of predicting hydrological responses at larger scales, such as the paddock to hill slope scales, all the way up to catchment scales (Blöschl and Sivapalan, 1995; Reggiani *et al.*, 1998). Most of the current generation of fully distributed physically-based models, such as SHE (Abbott *et al.*, 1986), fall into

the *upward* or *bottom-up* category. These models need an enormous amount of data on the physical characteristics and response of catchments and this limits their application.

In contrast, the *downward approach* develops a concept directly relevant to the space and time scale of interest (say catchment and annual) and gives a numerical representation to predict the catchment response. Success and failure of the conceptual model is tested through comparison of observations against predictions. Model complexity and/or improved process representations are added progressively to the model until it matches observations at finer scales of interest (say hill slope, and daily time scale). Manabe (1969) represented the annual water balance satisfactorily with a simple bucket representing soil water storage and an evaporative threshold. Milly and Dunne (1994) also showed that at continental and regional scales, simple lumped storage representation of surface hydrology was sufficient. However, at finer spatial and temporal scales, additional processes and complexities must be incorporated into the model (Sivapalan and Woods, 1995; Mohseni and Stefan, 1998).

The lumped conceptual models, an alternative to physically-based models, consider the catchment as a combination of interconnected conceptual stores without accounting for catchment geometry or spatial variability. Although the history of conceptual models is probably longer than that of physically based models, the *downward* or *top-down* approach was not followed in their early development. An example is the widely used AWBM model (Boughton, 1995) which divides total runoff between surface excess and baseflow. The model can be calibrated to reproduce catchment runoff response but cannot, explicitly, capture how a catchment responds dynamically to a change in land use. Sivapalan *et al.* (1996a,b) developed a more advanced conceptual model which partitions surface runoff between saturation excess and infiltration excess and then adds subsurface stormflow from an 'A' store that fills by infiltration and both loses and receives water from a deeper 'B' store that represents the groundwater system. An intermediate 'F' store is also included to conceptualise a connection between deep infiltration and recharge to the 'B' store. This store does not connect directly to the 'A' store. All three stores can lose water to the atmosphere by evaporation and transpiration. The resulting large scale catchment model (LASCAM) has been applied successfully to describe runoff from a number of catchments and can account for land use change, but still requires a minimum calibration of 22 model parameters and three initial store values (Sivapalan *et al.*, 2002).

More recently, the *downward approach* in model building has shown promising signs of success (Jothityangkoon *et al.*, 2001; Farmer *et al.*, 2003). Farmer *et al.* (2003) presented

a comparative analysis of the climate and landscape variability, which control the water balance of a number of diverse temperate and semi-arid Australian catchments. Jothityangkoon *et al.* (2001) systematically developed a water balance model, and applied it to a large semi-arid catchment in Western Australia. Similar models have also been developed and applied to other parts of the world (Atkinson *et al.*, 2002). All these studies show that the scale of interest, both time (annual to hourly) and space (point to ~1000 km<sup>2</sup>), and climate (aridity v. humidity) determine the required model complexity. Soil physical properties (such as depth and porosity) determine the storage capacity and ultimately play an important role in the formation of dynamic saturated areas for streamflow generation.

The recent efforts cited above have generally been devoted to water balance predictions of catchments that are in equilibrium. That means that, during the period of study, the flow generation processes of the catchment remain stationary and there are no changes in catchment physical attributes such as land use. However, it is well known that vegetation clearing alters the water balance of a catchment, as manifested for example by a rise of groundwater levels. Observations in a number of experimental catchments in the south-west of Western Australia show different rates of groundwater level rise, depending upon the extent and location of clearing (Schofield and Ruprecht, 1989). Streamflow generation processes of catchments have changed quite dramatically, once the rising groundwater levels reach the stream invert (Schofield and Ruprecht, 1989; Ruprecht and Schofield, 1991). The rising groundwater creates larger stream zone saturated areas and additional streamflow (Bari and Croton, 2000; Croton and Bari, 2001). The stream zone saturated areas stabilizes after 10 to 15 years of clearing and the catchment reaches a new equilibrium (Mauger *et al.*, 2001; Bari *et al.*, 2004). Analysing data from two sets of paired catchment experiments can determine the minimum model complexity required to capture the effect of land use change on catchment hydrological response on a monthly time scale.

This paper extends the work of Bari *et al.* (2004) which examined annual time steps. On an annual basis, it was shown that the response to land use change influenced the extent of the saturated area which controls the storm flow generation. However, at monthly time scale, there is a consistent lag between the peak monthly rainfall and runoff that cannot be explained without conceptualising a systematic gain/loss to storage content. To specify the framework of the conceptual model, data collected over 27 years from two sets of paired catchment experiments have been analysed. These data contain the key signals of the hydrological response to clearing. Then, a conceptual model

with additional complexity compared to the annual model, is developed to represent the monthly time step. The resulting model has four conceptual stores representing the flow generation processes. Ideally, the model should have the minimum number of parameters necessary to represent adequately the changes in flow generation processes resulting from land use changes. Unlike many previous conceptual models, these parameters are estimated *a priori* with minimal calibration.

## Description of the experimental catchments

The two sets of catchments were part of the five experimental catchments within the Collie River catchment (Fig. 1). The primary objective of the original experimental catchment programme was to quantify the effects of native forest clearing for agricultural development on streamflow and salinity. The area has a Mediterranean type climate, with cool, wet winters and warm to hot dry summers. The annual pan evaporation ranges from a maximum of 1600 mm to 1350 mm (Luke *et al.*, 1988). The soil profile typically consists of 50–650 cm highly permeable soil overlying 10–30 m of low permeability kaolinitic clay. The original vegetation was an open forest dominated by jarrah (*Eucalyptus marginata*), though marri, (*E. calophylla*) and wandoo (*E. wandoo*) to form an upper-storey to a height of 20–35 m (Bettenay *et al.*, 1980).

## DATA ANALYSES AND INTERPRETATION

To understand the changes in flow regime following clearing of native forest for pasture development, data from four experimental catchments (Fig. 1), monitored for three years before treatments, have been analysed.

### Wights and Salmon catchment

In the summer of 1976–77, the native forest of Wights was logged (Fig. 2). The cleared area was aerially sown to clover and grasses. The pre-clearing groundwater-induced saturated

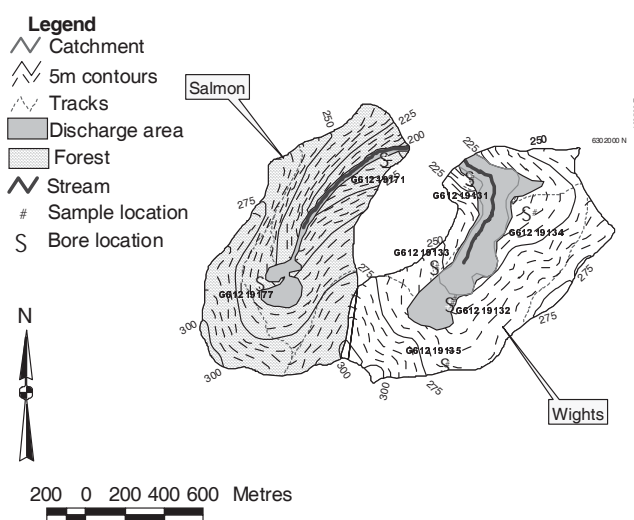


Fig. 2. Detail set up of Wights and Salmon catchments

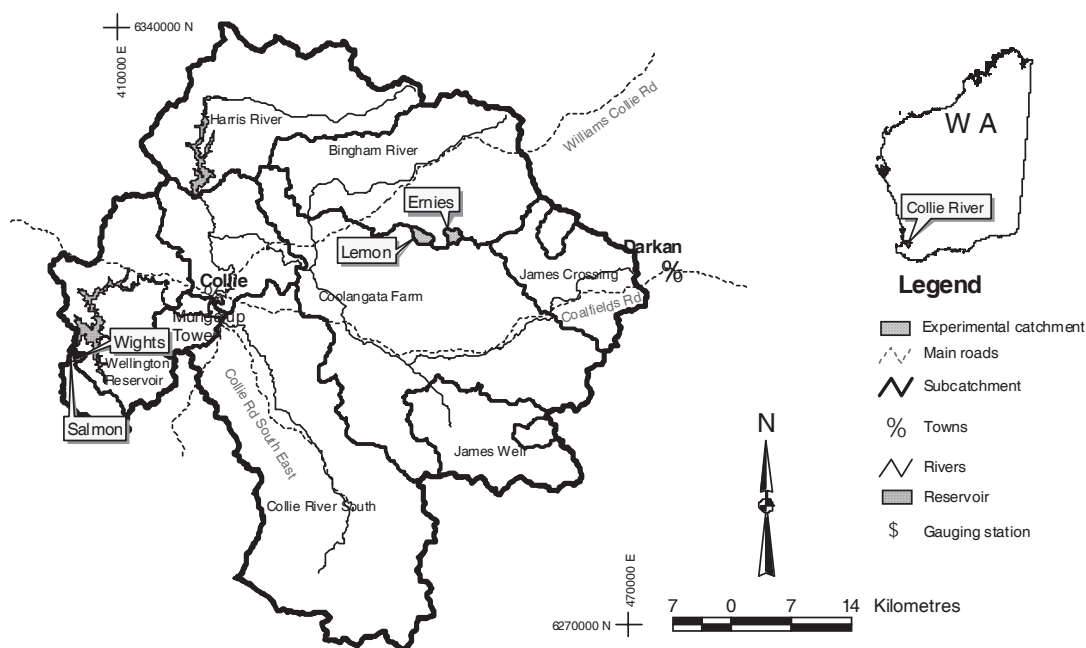


Fig. 1. Location of the experimental catchments

area at Salmon was 1% of the catchment area (Stokes, 1985). Most of the hydrographs for the piezometers at the Salmon catchment showed within year amplitudes of 1–4 m. The maxima were in October, at the end of the wet winter while minima were generally observed in May, at the end of a dry summer. The average depth to groundwater level varied spatially from nil to 15 m and there was a positive pressure head at the upper central part of the catchment (Fig. 3a). A bore located near the catchment divide showed the least response to rainfall, where the groundwater level was approximately 15 m below surface. The average lateral hydraulic conductivity of the aquifer is very low (Williamson *et al.*, 1987). From 1972–1998, the groundwater level in most of the piezometers showed a declining trend in Salmon catchment (Fig. 3a). Similar trends were observed in other control catchments in the south-west of Western Australia (Bari *et al.*, 1996). The exact cause of the declining trend is unknown but it may be associated with a prolonged period (1970–2000) of lower than average rainfall in Western Australia (Schofield and Ruprecht, 1989). During the pre-

clearing period, the groundwater levels of both catchments were well-correlated in annual amplitude and spatial distribution. Immediately following clearing at Wights catchment, there was an increasing trend in groundwater level with reduced within-year amplitude suggesting net recharge to the aquifer (Fig. 3b). The reduction in seasonal amplitude may result from the removal of native forest which previously transpired/extracted water from the aquifer. The relative increase in groundwater level ranged from 2–8 m depending upon the location of the bore in the landscape. However, by 1985, a new stability had been achieved and there was no change in seasonal amplitude or overall trend in groundwater level (Fig. 3b). Due to the increase in groundwater level, the stream zone saturated areas increased systematically and by 1985 had expanded to 18% of the catchment area (Fig. 2).

Measurement of porosity and soil moisture contents were also undertaken at the Wights and Salmon catchments. In 1973, five sites in each of the catchments were selected as representative of the hydrological provinces (Bettenay *et*

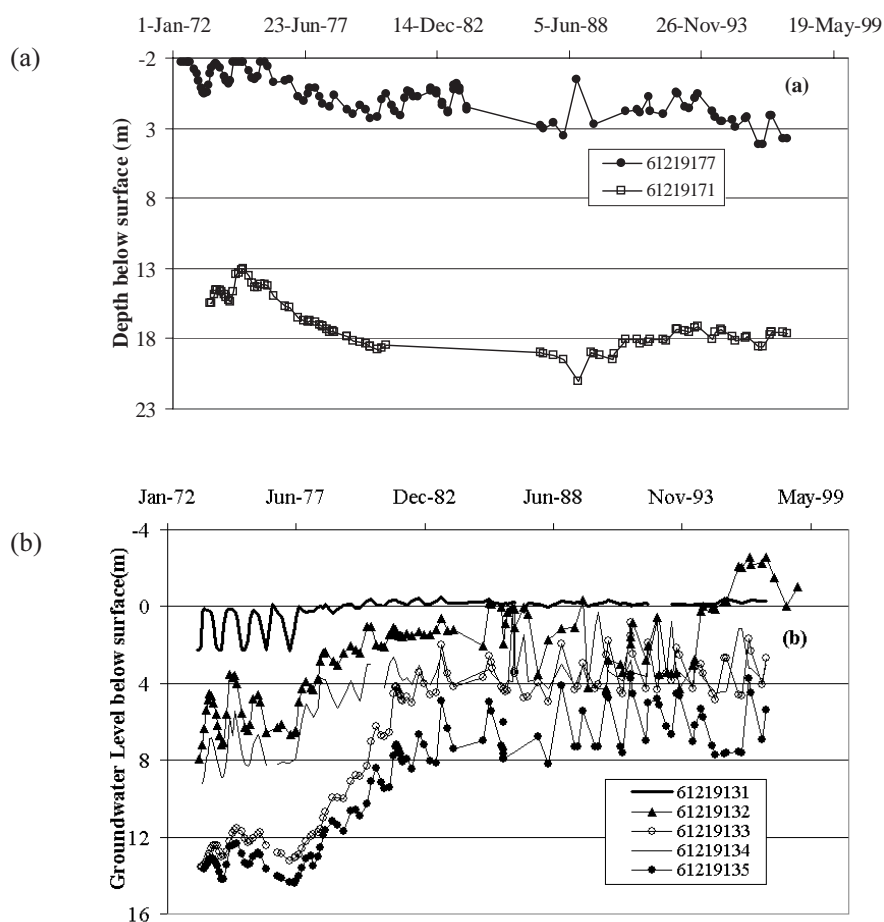


Fig. 3. Groundwater level variations observed at stream zones of (a) Salmon and (b) Wights catchments.



*al.*, 1980). The porosity of the Wights catchment is variable spatially and vertically. The average porosity of the top 3.5 m of the soil profile is  $0.375 \text{ m}^3 \text{ m}^{-3}$ , while in the 3.5–7 m depth interval soil porosity is the lowest,  $0.3 \text{ m}^3 \text{ m}^{-3}$ . The morphology of these sections of the profile has been described as surface soil and mottled zone respectively (Johnston, 1987). Below the mottled zone, the porosity of the underlying pallid zone increases to an average of  $0.425 \text{ m}^3 \text{ m}^{-3}$ , then reduces to approximately  $0.375 \text{ m}^3 \text{ m}^{-3}$ , particularly to the weathered zone of the profile. The soil moisture content generally increases with depth. The storage deficit was calculated; the difference between spatial average porosity and soil moisture content was calculated and was highest within the top 3.5 m depth of the soil profile. The storage deficit at the interface of surface soil and pallid zone of the soil profile was the lowest. As the soil moisture content generally increased with depth, the deficit in the deep clay profile was less than that of surface soil. On average, the storage deficit of the soil profile during the pre-clearing period at the Wights catchment was 1110 mm. The unsaturated storage change varied from one year to the next and, in the first three years after clearing, the average storage increase in the 6 m depth of the soil profile was some  $90 \text{ mm yr}^{-1}$  (Sharma *et al.* 1987a).

Following clearing, annual stream flow increased systematically until a new equilibrium between the groundwater recharge and discharge was achieved by 1985. The average annual increase in streamflow, after stability, is in the order of  $300 \text{ mm yr}^{-1}$ . The increase in streamflow was due in part to spatial and temporal enlargement of a perched groundwater system and deep groundwater discharge zones (Bari *et al.*, 1996; Ruprecht and Stoneman, 1993; Bari *et al.*, 2004).

#### *Ernies and Lemon catchments*

In the summer of 1976–77, merchantable timber was logged from the lower 53% of the Lemon catchment (Fig. 4). The cleared area was sown to clover-based pastures to be used for grazing sheep. Ernies catchment was established as a control. Most of the groundwater observation bores at Ernies catchment were installed just upstream of the gauging station, perpendicular to the stream line. Most of the bores were installed in groups of two, measuring both the shallow intermittent and deep permanent groundwater levels. The depths of the shallow and deep bores ranged from 1 to 3 m and 12 to 36 m respectively. At Ernies catchment, the permanent groundwater level was about 15–20 m below the stream invert (Fig. 5) and did not contribute to streamflow generation. The seasonal amplitude of the groundwater level was much less than that of Wights and Salmon, indicating less recharge to and loss from the aquifer. There was a

general decline in groundwater level (Fig. 5a). The shallow bores were dry in most of the dry months of the year (October–May). In the winter months, when heavy rainfall occurs in May to June, a shallow intermittent groundwater system developed on cap rock or clay (perched seasonal groundwater), which is evident in the shallow bores (Fig. 5a). Therefore, part of the stream zone becomes saturated in winter months and dries out in summer. The development of intermittent groundwater level plays a very important role in the streamflow generation.

At the Lemon catchment, groundwater observation bores were established in the valley, mid-slope and up-slope areas. During the pre-treatment period, the depths to groundwater level and the seasonal amplitude were very similar for both catchments (Fig. 5). The deep, permanent groundwater system beneath Lemon catchment started to rise following clearing, both in the valley and upslope cleared areas (Fig. 5b). The seasonal amplitude virtually disappeared for the period 1978–86, indicating the low transmissivity of the aquifer and there was no groundwater discharge to the stream. The rate of groundwater level rise increased with time, intersected the surface by 1987 (Fig. 5b), and may have achieved a new stability by 1996. Some of the bores located in the stream zone, recorded more than 6 m positive piezometric pressure above the soil surface by 1998. Part of the stream zone became ‘permanently’ saturated which is supported by the presence of groundwater in the shallow bores throughout the year. In 1996, the groundwater induced saturated area was much lower than that of Wights catchment.

As for the Wights and Salmon catchments, soil moisture content and porosity measurements were undertaken in the Lemon and Ernies catchments (Bettenay, 1980; Johnston, 1987). The catchment average porosity of the top soil of Lemon catchment was  $0.3 \text{ m}^3 \text{ m}^{-3}$ , slightly less than that of Wights. Porosity increased with depth and in the deep-clay profile. Soil moisture content of Lemon catchment was generally lower than that of Wights. The moisture content of the profile increased with depth. The storage deficit was much higher, particularly in the deep clay section of the profile. On average, the deficit is 2390 mm over the full depth of the profile. Within two years of clearing of native forest from Lemon catchment, there was a significant increase in unsaturated soil water storage (Sharma *et al.*, 1987a).

There was an immediate increase in annual flow volume of approximately 25 mm from Lemon catchment following clearing. When the groundwater reached the soil surface in 1987, there was a second phase of significant change in streamflow and runoff generation process; average annual streamflow increased by 100 mm from 1993–1998.

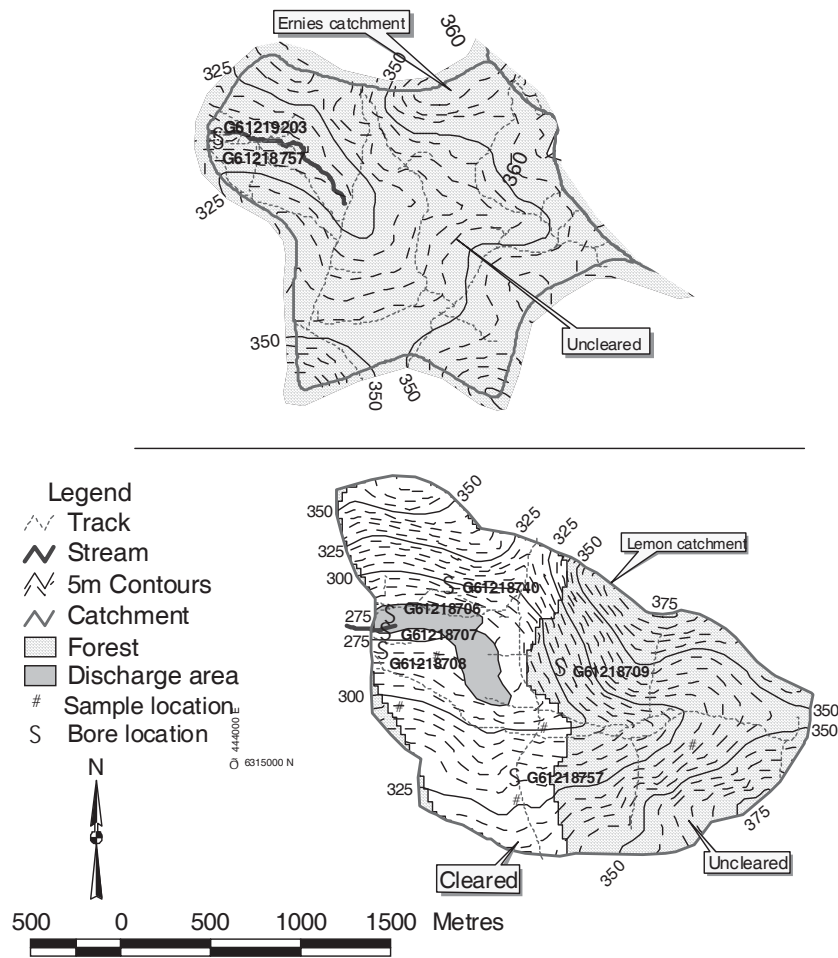


Fig. 4. Detail set-up of Ernies and Lemon catchments

Measurement from satellite photographs revealed that stream zone saturated areas increased from zero in 1987 to approximately 8% of the catchment by 1996 (Bari *et al.*, 2004).

#### MONTHLY WATER BALANCE MODEL

Bari *et al.* (2004) recognised the importance of the stream zone saturated area in controlling annual runoff. However, analyses and interpretation of the data in relation to monthly hydrological processes suggest that four inter-connecting stores are required to represent the landscape. Figure 6 shows the conceptual representation of a hill slope of a hypothetical catchment. The stores for the monthly water balance model are: (i) Upper Store, (ii) Subsurface Store, (iii) Groundwater Store, and (iv) Stream zone Store. The catchment is represented by the ‘open-book’ approach (Fig. 6). Average hill slope length ( $L_s$ ) is calculated from catchment area ( $A_t$ ) and stream length ( $L$ ) as:

$$L_s = \frac{A_t}{2L} \quad (1)$$

If a catchment is not cleared for agriculture, then the groundwater may lie below the stream bed (Fig. 6b) and a loss term would suffice. However, the store is required to capture groundwater rise after clearing. The Upper Store generates surface runoff ( $Q_{r1}$ ) and interflow ( $Q_2$ ). Water percolates ( $I$ ) from the Upper Store to the deep unsaturated Subsurface Store (Fig. 6c). Most of the percolated water is transpired back to the atmosphere from the Subsurface Store by deep-rooted trees. The remaining deep percolation reaches the deep groundwater store ( $R_p$ ), and is responsible for the changes of groundwater level as observed in all four catchments (Fig. 3, Fig. 5). When the groundwater level rises following clearing and reaches the stream bed, the stream zone store is created (Fig. 6c). The groundwater discharges ( $Q_b$ ) to the stream through the stream zone store. Additional surface runoff ( $Q_{r2}$ ), is generated from the ‘impervious’ stream zone saturated area. The model differs in concept to LASCAM (Sivapalan *et al.*, 1996 a,b) in that the groundwater store is linked to the stream via the stream zone store once the groundwater reaches the stream bed.

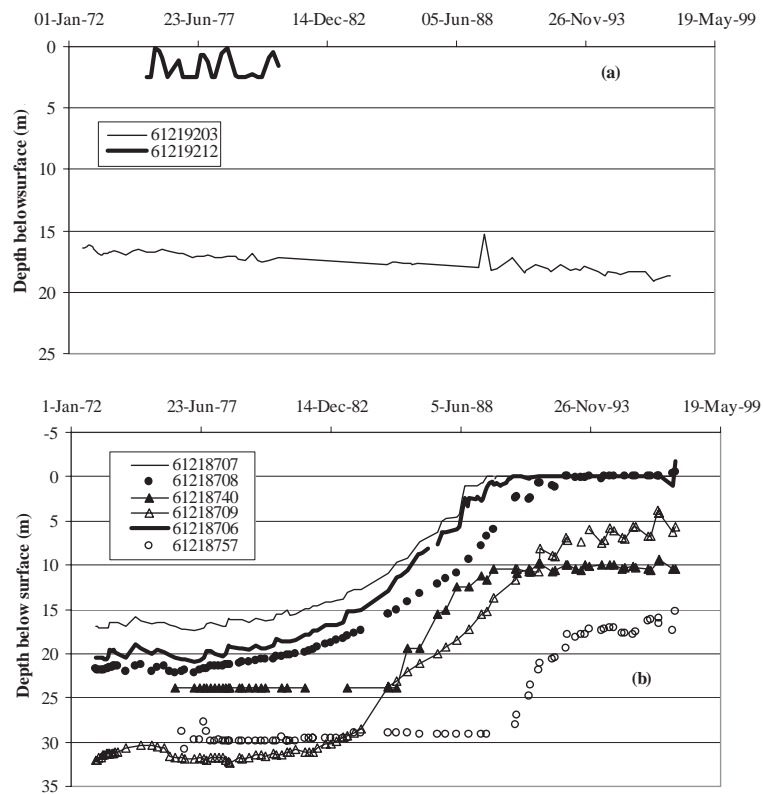


Fig. 5. Groundwater level variations observed at (a) Ernies and (b) Lemon catchment

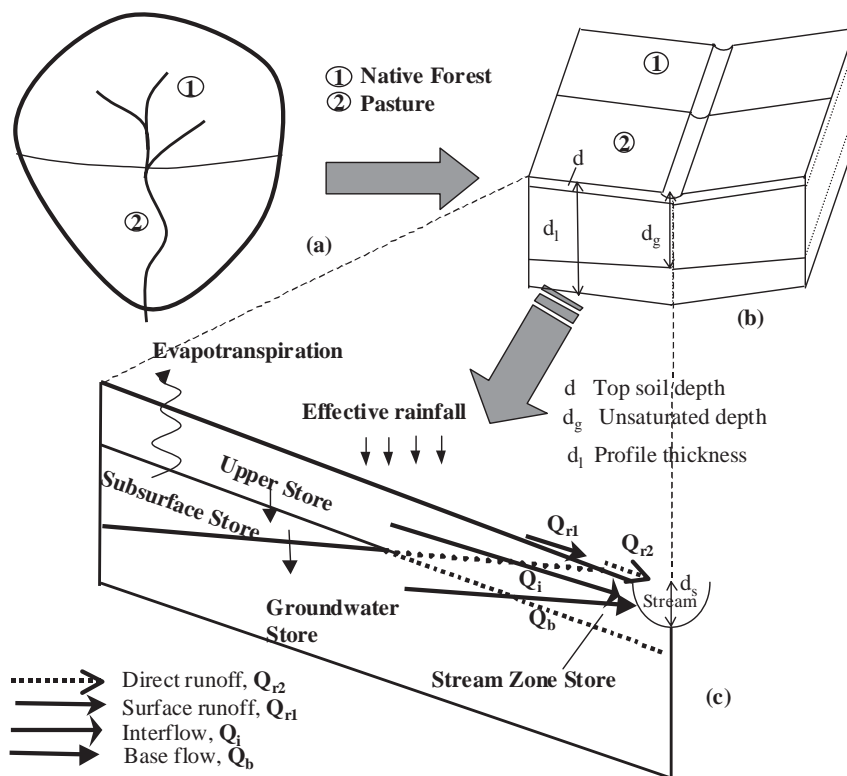


Fig. 6. Schematic representation of a hill slope by four-store model



Parameters are sought *a priori* and calibration requirements are minimised. The water balance of the four stores is described next (see Appendix A for explanation of symbols used).

#### Evapotranspiration

Evapotranspiration is the most important component of the water balance of these four catchments. It has three components: (i) interception, (ii) plant transpiration and (iii) soil evaporation. Rainfall interception depends upon the Leaf Area Index ( $LAI$ ) and the interception storage of the forest canopy. Rainfall interception ( $I_a$ ) on the plant canopy is modelled using an empirical equation:

$$I_a = a_2 R \quad (2)$$

Effective rainfall ( $RE$ ) passes through the plant canopy and becomes available for infiltration into the ground:

$$RE(t+1) = R(t, t+1) - I_a(t+1) \quad (3)$$

Evaporation of water intercepted by the plant canopies has the first call on potential evapotranspiration. Residual potential evapotranspiration ( $RET$ ) is the energy available for plant transpiration and soil evaporation processes.

$$RET = PET - I_a \quad (4)$$

#### Soil evaporation

Actual soil evaporation depends on atmospheric demand, soil moisture and canopy shading. Soil evaporation takes place from upper zone ( $E_{su}$ ) and from stream zone store ( $E_{ss}$ ), if it exists.

$$E_{su} = RET \frac{W}{W_{mx}} e^{-csLAI} \quad (5a)$$

$$E_{ss} = RET \frac{W_{mx}}{W_{mx}} e^{-csLAI} \quad (5b)$$

The residual potential evapotranspiration ( $RET$ ) is reduced to:

$$RET = RET - E_{su} - E_{ss} \quad (6)$$

#### Plant transpiration

Roots of the native jarrah forest often reach the groundwater table. In addition to the Upper Store, deep-rooted plants extract water from Subsurface Store and groundwater for transpiration (Carbon *et al.*, 1980; Carbon *et al.*, 1981). The rooting depth of pasture is likely to be limited to the surface soil only, which is only few metres deep (Ward *et al.*, 2002). Non-linearity between plant transpiration and soil moisture

content was introduced as found by Wood *et al.* (1992). Actual plant transpiration involves physical and physiological processes and is the most difficult component to quantify. It has been assumed that plants extract water from all stores depending on available energy, soil moisture condition, Leaf Area Index and presence of root volume. Actual plant transpiration from the upper zone store ( $E_{tu}$ ) is:

$$E_{tu} = \alpha_t RET \frac{RT_u}{RT_t} \left[ 1 - \left( 1 - \frac{W}{W_{mx}} \right)^{tu} \right] \frac{LAI}{LAI_{mx}} \quad (7)$$

When the transpiration from upper zone stores are completed, the residual potential evapotranspiration becomes:

$$RET = RET - E_{tu} \quad (8)$$

Transpiration from the sub-surface store ( $E_{tl}$ ) and groundwater store ( $E_{tg}$ ) are calculated as:

$$E_{tl} = \alpha_t RET \frac{RT_l}{RT_t} \left[ 1 - \left( 1 - \frac{W_l}{W_{lmx}} \right)^{tl} \right] \frac{LAI}{LAI_{mx}} \quad (9)$$

$$RET = RET - E_{tl} \quad (10)$$

$$E_{tg} = \alpha_t RET \frac{RT_g}{RT_t} \frac{LAI}{LAI_{mx}} \quad (11)$$

Therefore, after transpiration from the groundwater store, the residual potential evapotranspiration is:

$$RET = RET - E_{tg} \quad (12)$$

#### Upper Store

The Upper Store represents the highly conductive surface soil, generally a few metres deep. The maximum capacity of the store is dependent upon the catchment average soil depth of the topsoil ( $d$ ) and the porosity ( $\phi_u$ ). The maximum capacity ( $W_{mx}$ ) and water content ( $W$ ) can be represented as:

$$W_{mx} = d\phi_u \quad (13a)$$

$$W = d\theta_u \quad (13b)$$

When the water content of the store exceeds its maximum capacity, saturation excess runoff occurs:

$$Q_{r1} = W - W_{mx} \quad \text{if } W > W_{mx} \quad (14a)$$

$$Q_{r1} = 0.0 \quad \text{if } W < W_{mx} \quad (14b)$$

The second component of surface runoff ( $Q_{r2}$ ) is generated from the groundwater induced stream zone saturated area. This follows the observation of Bari *et al.* (2004) that the stream zone saturated area increases to a new equilibrium some time after clearing and becomes an important determinant of the annual water balance. Therefore, the total surface runoff ( $Q_r$ ) is:

$$Q_r = Q_{r1} + Q_{r2} \quad (15)$$

The Upper Store generates interflow if the moisture content exceeds a threshold. It is formulated based on lateral conductivity of the top layer ( $K_{ul}$ ) water content and a threshold parameter ( $W_i$ ), such as:

$$Q_i = 0 \quad \text{if } W < W_i \quad (16a)$$

$$Q_i = K_{ul} \left( \frac{W - W_i}{W_{mx} - W_i} \right)^{ia} \quad \text{if } W > W_i \quad (16b)$$

Percolation ( $I$ ) is defined as the rate of vertical drainage from Upper to Lower unsaturated Store and is dependant upon the vertical conductivity ( $K_{uv}$ ) and water content ( $W$ ) of the Upper Store. Survey of the relevant literature shows that the moisture content of the Subsurface Store ( $W_l$ ) plays an important role in controlling percolation, in the form of preferential flow (Johnston, 1987). Percolation from Upper Store is formulated as:

$$I = K_{uv} \left[ 1 + pb \left( 1 - \frac{W_l}{W_{mx}} \right)^{pa} \right] \left( \frac{W}{W_{mx}} \right) \quad (17)$$

The water balance of the Upper Store at any particular time can be expressed as:

$$W(t+1) = W(t) + RE(t, t+1) - E_{lu}(t, t+1) - E_{su}(t, t+1) - Q_{r1}(t, t+1) - Q_i(t, t+1) - I(t, t+1) \quad (18)$$

#### Subsurface Store

The Subsurface Store represents the deep unsaturated soil profile below the surface soil and above the permanent groundwater table (Fig. 6). It receives water percolated from upper store and loses to the groundwater system due to recharge. This store also loses/gains to/from the groundwater store, due to rise/fall of the groundwater level. The store content depends upon the catchment-wide average depth of the unsaturated profile below the surface soil ( $d'_g - d$ ) and moisture content ( $\theta_l$ ), which can be expressed as:

$$W_{lmx} = (d'_g - d)\theta_l \quad (19a)$$

$$W_l = (d'_g - d)\theta_l \quad (19b)$$

$$W_{ld} = (d'_g - d)\theta_f \quad (19c)$$

Transpiration from this store is similar to the Upper Store (Eqn. 9). Recharge ( $RI$ ) to the Groundwater Store occurs when the water content exceeds field capacity ( $W_{ld}$ ):

$$RI = 0 \quad \text{if } W_l < W_{ld} \quad (20a)$$

$$RI = K_{lv} \left( \frac{W_l - W_{ld}}{W_{lmx} - W_{ld}} \right)^{3.3}, \quad \text{if } W_l > W_{ld} \quad (20b)$$

The groundwater level changes due to recharge ( $RI$ ), base flow ( $Q_b$ ) and transpiration ( $E_{tg}$ ). The changes in water fluxes and groundwater level below the subsurface store are:

$$\Delta W_l(t+1) = RI(t, t+1) - Q_b(t, t+1) - E_{tg}(t, t+1) \quad (21)$$

$$\Delta d_g = \frac{\Delta W_l (d'_g - d)}{[W_{lmx} - W_l]} \quad (22)$$

Therefore, the depth to groundwater level is updated as:

$$d_g(t+1) = d_g(t) + \Delta d_g(t, t+1) \quad (23)$$

When the groundwater level fluctuates, the Sub-surface and Groundwater stores exchange part of the water content following:

$$\Delta W_{gl} = \left( \frac{\Delta d_g}{d'_g - d} \right) W_l \quad \text{if } \Delta d_g < 0 \quad (24a)$$

$$\Delta W_{gl} = \left( \frac{\Delta d_g}{d_l - d'_g} \right) W_g \quad \text{if } \Delta d_g > 0 \quad (24b)$$

Therefore, the subsurface store content at time ( $t+1$ ) is:

$$W_l(t+1) = W_l(t) + I(t, t+1) - E_{tl}(t, t+1) - RI(t, t+1) + \Delta W_{gl}(t, t+1) \quad (25)$$

#### Groundwater Store

The conceptual Groundwater Store represents the water balance of the permanent groundwater system (Fig. 6). The water content ( $W_g$ ) can be expressed as:

$$W_g = (d_l - d'_g)\theta_l \quad (26)$$

If the plant root reaches the Groundwater Store and there is any residual transpiration potential, then the plants can transpire groundwater (Eqn. 11). The gradient of the permanent groundwater system ( $\beta$ ) is considered to be half

of the catchment-wide average slope of the ground surface (Fig. 6). As the catchment is represented by an ‘open-book’ approach, the baseflow to the stream can be calculated as:

$$Q_b = K_{ll} L |(d_s - d_g)| \tan \beta \quad \text{if } d_g < d_s \quad (27a)$$

$$Q_b = 0 \quad \text{if } d_g > d_s \quad (27b)$$

Therefore, the groundwater storage is updated as:

$$W_g(t+1) = W_g(t) + Rl(t, t+1) - Q_b(t, t+1) - \Delta W_{gl}(t, t+1) - E_{tg}(t, t+1) \quad (28)$$

#### Stream zone store

This store is transient, and is created by the deep groundwater system only. If the groundwater level is not on or above the stream invert, this store ceases to exist. It covers part of the Upper store and is considered as ‘impervious’ ( $A_i$ ). Therefore:

$$Q_{r2} = 0 \quad \text{if } d_g > 0 \quad (29a)$$

$$Q_{r2} = \frac{A_i}{A_t} RE \quad \text{if } d_g < 0 \quad (29b)$$

When the groundwater level increases, the stream zone saturated area expands ( $\Delta A_i$ ) and covers part of the upper store and vice versa. It can be calculated as:

$$\Delta W_{sg} = \left( \frac{\Delta A_i}{A_t} \right) W \quad \text{if } d_g < 0 \quad (30)$$

The stream zone store water content at any time ( $t+1$ ) is then:

$$W_{sg}(t+1) = W_{sg}(t) + \Delta W_{sg}(t, t+1) \quad (31)$$

#### Streamflow

Total streamflow is the sum of surface runoff, interflow and baseflow components:

$$Q_t = Q_r + Q_i + Q_b \quad (32)$$

#### MODEL APPLICATION

The model was applied to all four catchments for the whole period of study (1974–1998). For the Ernies and Salmon control catchments, the first five years of data were used for calibration. As there were significant changes in flow generation process in the Wights and Lemon catchments, monthly streamflow data up to 1985 and 1987 respectively

were used for calibration. The rest of the streamflow data were used for model verification. Monthly pan evaporation was calculated from annual values (Luke *et al.*, 1988) using a simple harmonic function. The interception parameter ( $a_2$ ) was adjusted to approximately 13% of rainfall, which is typical for Western Australian jarrah forest (Williamson *et al.*, 1987; Croton and Norton, 1998). For pasture interception, only the Leaf Area Index was changed and the predicted interception was similar to the results obtained from modelling grassland pasture in the Swan Coastal plain of Western Australia (Raper and Sharma, 1989). The relative root volume of jarrah (*E. marginata*) forest and pasture in the upper and lower stores was estimated from the literature. Roots of the jarrah often reach the groundwater table and extract water for transpiration (Carbon *et al.*, 1980; Carbon *et al.*, 1981; Sharma *et al.*, 1987a). It was estimated that the relative root volumes in the upper, subsurface and groundwater stores are approximately 60%, 35% and 5%, respectively. Mean Leaf Area Index of native forest and pasture was obtained from previous studies (Mauger, 1988; Bari and Croton, 2000; Croton and Bari, 2001).

The average depth of the topsoil ( $d = 2.5m$ ), average porosity ( $\phi_u, \phi_l = 0.4$ ), profile thickness ( $d_l$ ) and depth to groundwater level below stream ( $d_g$ ) were estimated from drilling information and fixed *a priori*. Catchment-average surface slope, stream depth ( $d_s$ ) and stream length ( $L$ ) were determined from topographic maps and site visits and again fixed *a priori*. The catchment average calibrated lateral hydraulic conductivity ( $K_{ll}$ ) of the top-soil ranges from 380 to 580 mm month<sup>-1</sup>, much lower than the experimental evidence of 0.2 to 22.7 m day<sup>-1</sup> (Sharma *et al.*, 1987b). Recent application of a fully distributed model to Lemon catchment shows that the lateral conductivity of the top layer needs to be calibrated to 15 m day<sup>-1</sup> (Croton and Bari, 2001). The apparent differences in lateral conductivity require further investigation. The vertical ( $K_{lv}$ ) and lateral ( $K_{lu}$ ) conductivity of the clay layer was calibrated to 17 mm month<sup>-1</sup> and 9.53–15.53 mm month<sup>-1</sup> respectively (Table 1). These are less than slug test results of 69 to 228 mm month<sup>-1</sup> (Peck and Williamson, 1987). One possible explanation is that the values used here represent the catchment average over a monthly time step. At present therefore, it is not possible to avoid obtaining estimated values by matching the observed and predicted hydrographs. One parameter ( $ia$ ) represents the non-linear relationship between moisture content and conductivity (Eqn. 16) and another two ( $pb$ ,  $pa$ ) represent the leakage from upper to lower store (Eqn. 17). Once calibrated to one catchment, these parameters remained unchanged across all other catchments in the study (Table 1).

Table 1. Parameter values showing best fit for all catchments

Parameter	Unit	Salmon	Wights	Ernies	Lemon
$\alpha_t$	(-)	1.95	2.0	1.75	1.9
$K_{ul}$	mm month <sup>-1</sup>	390	580	380	380
$ia$	(-)	3.15	3.15	3.15	3.15
$K_{uv}$	mm month <sup>-1</sup>	107	57	87	87
$K_{ll}$	mm month <sup>-1</sup>	15.53	9.53	15.53	15.53
$pb$	(-)	0.5	0.5	0.5	0.5
$pa$	(-)	1.5	1.5	1.5	1.5
$K_{lv}$	mm month <sup>-1</sup>	17.0	17.0	17.0	17.0

## Results

### GROUNDWATER SYSTEM

The groundwater level predicted at Wights catchment remained stable during the pre-clearing period and then rose steadily with time as did the measured values (Fig. 7a). In contrast to the observations, the predicted groundwater level at the Salmon catchment showed a slight increase. This contradictory result may be related to the initial condition of the model (Fig. 7b). At Lemon catchment, an initial depth to groundwater level of 15 m was adopted for model calibration. Similar trends in the observed and predicted groundwater levels were evident (Fig. 7c). The initial depth

to groundwater level at Ernies catchment was assumed to be 15 m. To achieve stability, the initial level was changed a few times and the final predicted conceptual groundwater stabilised at about 6 m below the stream bed (Fig. 7d).

### MONTHLY STREAMFLOW

The observed and predicted monthly streamflow matched reasonably well for the Wights catchment. The Wights catchment was cleared in 1977, one of the lowest flow years in record. The predicted monthly flow during the dry period of the year (January to May) was higher than the observed (Fig. 8a). The predicted and observed streamflows in August were 62 mm and 82 mm respectively. The model also over-predicted the flow for September, November and December. Between the period of clearing and achieving new stability, the highest annual streamflow was in 1983. The predicted flow matched well for the low-flow period of the year. During June, July and September, there was an under prediction (Fig. 8b). The lowest flow was in 1994 after the catchment reached a new stability. The model under-predicted the flow, except for August and September (Fig. 8c). In 1996, streamflow was the highest. The model prediction was excellent from January to May. Except for July and November, the predicted flow was lower than observed for the rest of the year (Fig. 8d).

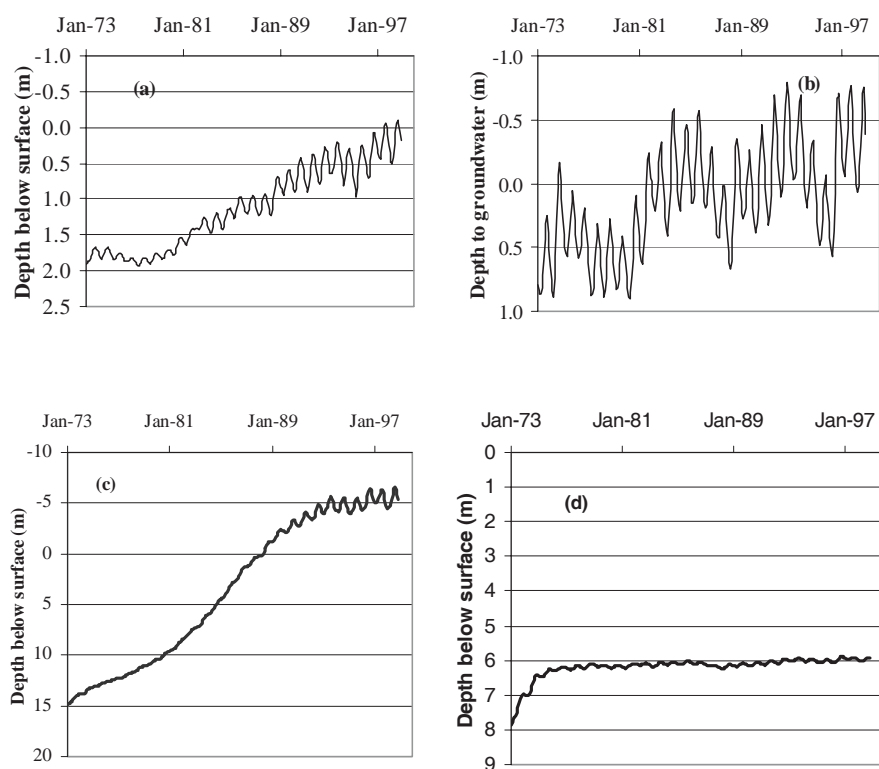


Fig. 7. Conceptual groundwater level at (a) Wights, (b) Salmon, (c) Lemon and (d) Ernies catchments

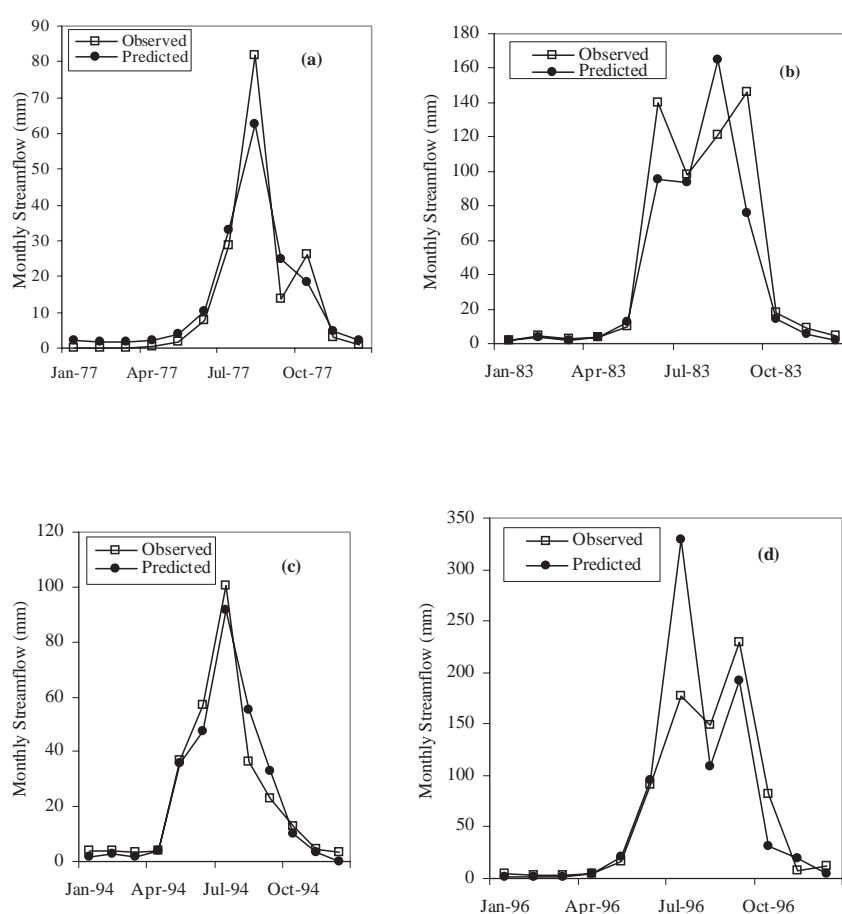


Fig. 8. Modelled and observed monthly streamflow at Wights catchment for (a) 1977, (b) 1983, (c) 1994 and (d) 1996

At Salmon catchment, 1980 was one of the low-flow years and there was no predicted or recorded flow during January to April. However, from May to July, the predicted streamflow was slightly higher than observed. The model predicted 47 mm for August while the observed flow was 56 mm (Fig. 9a). The predicted streamflow for the rest of the year matched well except for October. The catchment received the lowest annual rainfall in 1987. The predicted streamflow was higher than the observed except for the months when the stream was not flowing (Fig. 9b). After two successive years of low flow, the streamflow observed in 1988 was one of the highest following annual rainfall of 1425 mm. The predicted flows for June and July were higher and for September and October lower than observed (Fig. 9c). The predicted streamflow for 1996 was higher than observed during the onset of the wetting period, June–July. In fact the predicted flow for July was 197 mm, more than double of the observed flow of 89 mm (Fig. 9d).

At Lemon catchment there was a good match between the observed and predicted monthly streamflow. During the wetting phase, May–June, the predicted flow was higher

than observed but matched the highest recorded flow of 1976 very well (Fig. 10a). In 1982, Lemon catchment produced an average annual streamflow and this was predicted well, except for January and February (Fig. 10b). In January 243 mm of rain fell in three days and the catchment generated 18 mm of runoff, but the model predicted only 4 mm. In 1991, Lemon catchment flowed over the whole year as the groundwater system reached the streambed and created ‘permanent’ saturated areas (Fig. 5b). The model gave satisfactory predictions of streamflow, but recessions were higher than observed (Fig. 10c). The predicted streamflow was generally well-matched for 1998, when the groundwater system stabilised (Fig. 10d).

The observed streamflow was the lowest at Ernies catchment. In 1974, it generated 72 mm runoff, the highest in the record. The model predicted 49 mm due mainly to the lower predictions for July and August (Fig. 11a). The predicted onset of streamflow was earlier than observed, for an average flow-year of 1981. In 1998 the predicted onset of streamflow was higher and the predicted recession was lower than observed (Fig. 11c). In 1990, the prediction

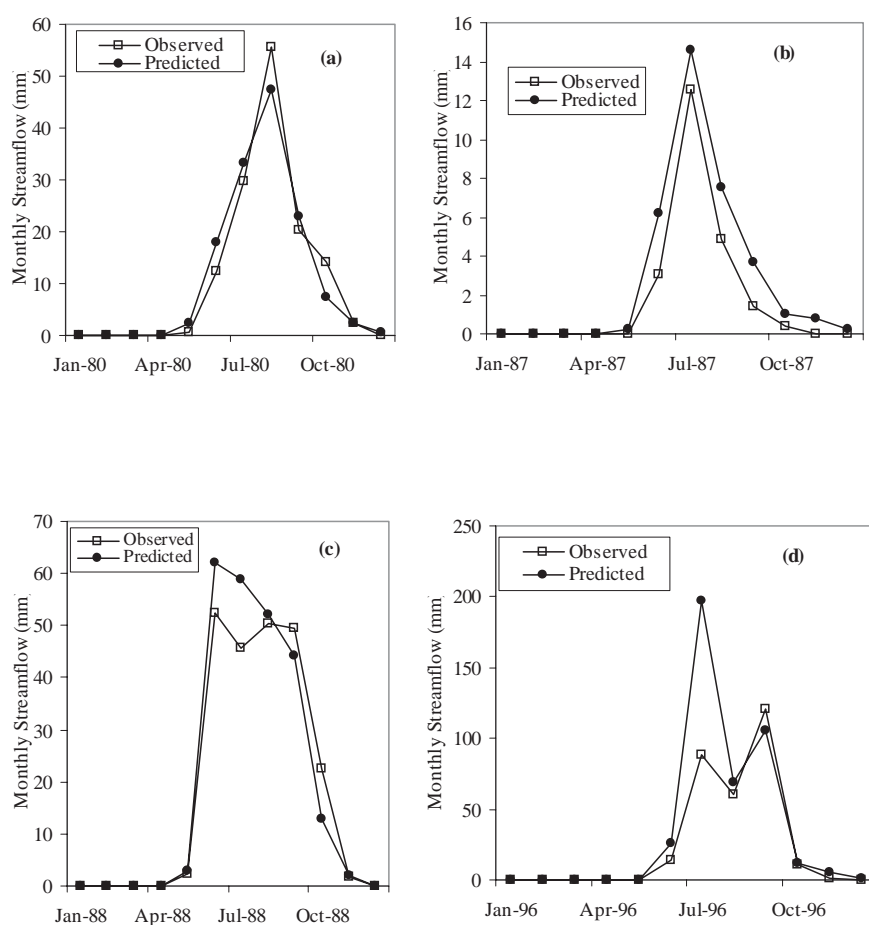


Fig. 9. Modelled and observed monthly streamflow at Salmon catchment for (a) 1980, (b) 1987, (c) 1988 and (d) 1996

was lower than observed except for July (Fig. 11d).

Relationships between the observed and modelled streamflow for all four catchments gave coefficients of determination ( $R^2$ ) of 0.68 to 0.88 (Fig. 12). At the Wights and Salmon catchments, the streamflow was grossly over-predicted for some months. At Ernies catchment, the lowest  $R^2$  was obtained because the model often predicted flow ranging up to 5 mm, when the stream was not flowing at all (Fig. 12d).

#### ANNUAL STREAMFLOW

During the pre-treatment period (1974–76), the annual streamflow at the Wights catchment was over-predicted. The model under-predicted the annual streamflow during 1981–87 when the groundwater system was rising and reaching a new equilibrium. The model over-predicted 6 out of 25 years at Wights catchment, with an overall water balance error of –7.8% (Table 2). At Salmon catchment, the observed and predicted streamflow for the period 1974–98 totalled 2964 mm and 2966 mm respectively, with a coefficient of

determination ( $R^2$ ) of 0.9 (Table 2). In 1974 the annual rainfall was 1456 mm and the catchment yielded 365 mm of streamflow. The observed streamflow is less than that of 1996 when the catchment received an annual rainfall of 1686 mm and yielded a runoff of 296 mm.

The observed and predicted annual streamflow at the Lemon catchment were well matched during the period 1974–87. The predicted annual total was less than observed for 1982 and also less for the highest flow year of 1996.

Table 2. Observed and predicted streamflow for the period 1974–1998

Catchment	Observed flow(mm)	Predicted flow(mm)	Error (%)	Coefficient of Determination
Wights	9138	9033	-7.8	0.90
Salmon	2964	2966	0.1	0.90
Lemon	1469	1466	-0.3	0.95
Ernies	212	227	7.4	0.83



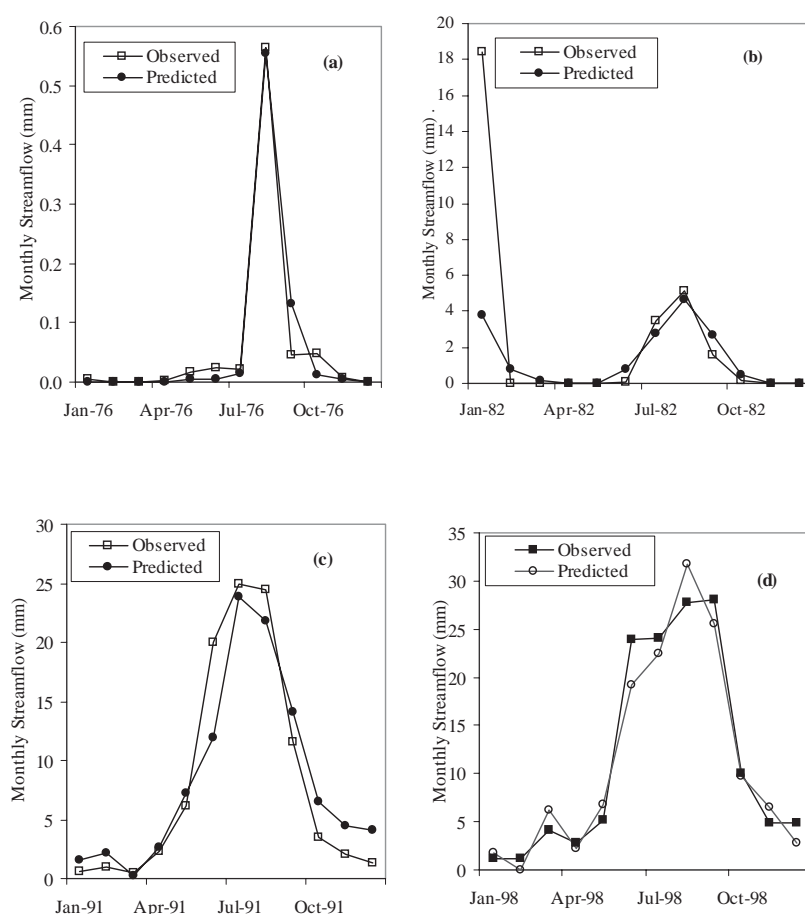


Fig. 10. Modelled and observed monthly streamflow at Lemon catchment for (a) 1976, (b) 1982, (c) 1991 and (d) 1998

Overall, the predicted stream yield was 1466 mm, 0.3% less than observed (Table 2). The observed annual streamflow at the Ernies catchment in 1974 was 72 mm, much higher than the Lemon catchment. Except for 1974, the predicted annual streamflow for the other high-flow years is reasonably matched. For some of the low-flow years the predicted annual streamflow was higher than observed but other years are well matched. Overall the observed and predicted sums were 212 mm and 227 mm respectively (Table 2).

#### WATER BALANCE COMPONENTS

The annual rainfall at the Salmon catchment is about 15% higher than that of the Wights catchment, while at the Ernies and Lemon catchments, monthly rainfall was estimated to be identical. At the Ernies and Salmon catchments, total interception during 1974–98 was 2314 mm and 3618 mm respectively (Table 3). Similar results were also obtained from interception experiments conducted in the jarrah forest of Western Australia (Williamson *et al.*, 1987; Croton and

Norton, 1998). Following clearing at the Lemon and Wights catchments, both the transpiration and interception decreased and the soil evaporation increased (Table 3). The soil evaporation was 11.5% of rainfall at the Wights catchment and 14% of rainfall in the cleared area of Lemon catchment, similar to the grassland pasture site in the Swan Coastal Plain of Western Australia (Raper and Sharma, 1989).

The water content of all the four stores at the Salmon catchment increased at the end of the simulation (Table 3). The largest increase was in the groundwater store, due to a slight rise in the predicted groundwater level (Fig. 7b). The predicted stream zone saturated area had a within year variation from nil to 5% of the catchment area (Fig. 13a), similar to the estimated mean of 2% (Stokes, 1985; Bari *et al.*, 2004). The predicted soil moisture content in the lower unsaturated store increased slightly, due to increase in store content (Fig. 14). At the Wights catchment, water volume in all four stores increased, with the largest increase being in the subsurface unsaturated store (Table 3). The predicted soil moisture content, before clearing, was slightly less than

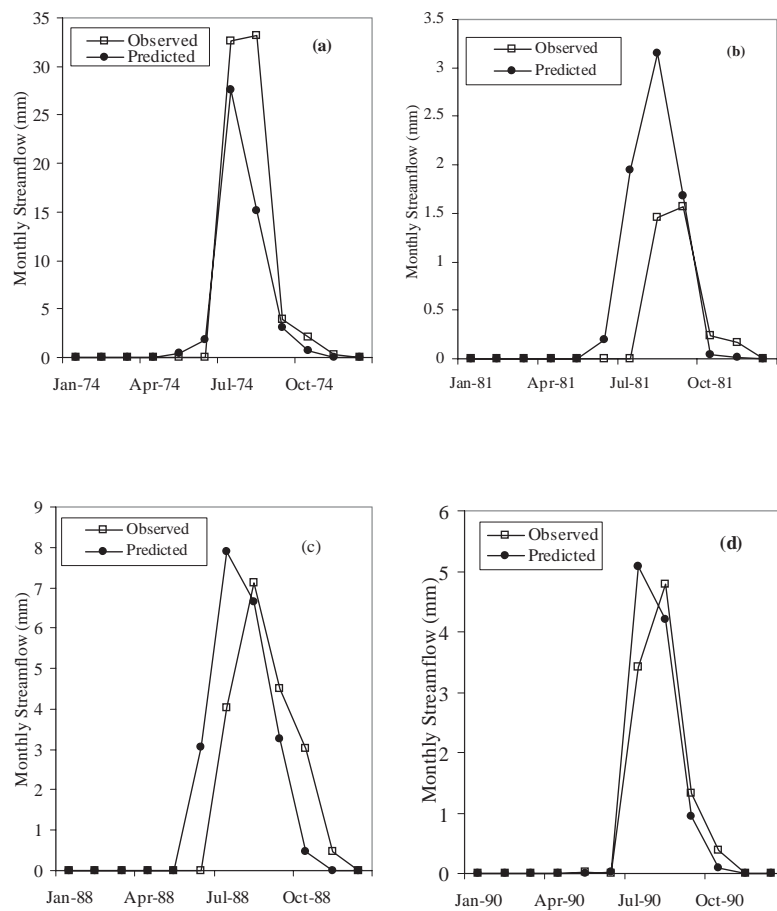


Fig. 11. Modelled and observed monthly streamflow at Ernies catchment for (a) 1974, (b) 1981, (c) 1988 and (d) 1990

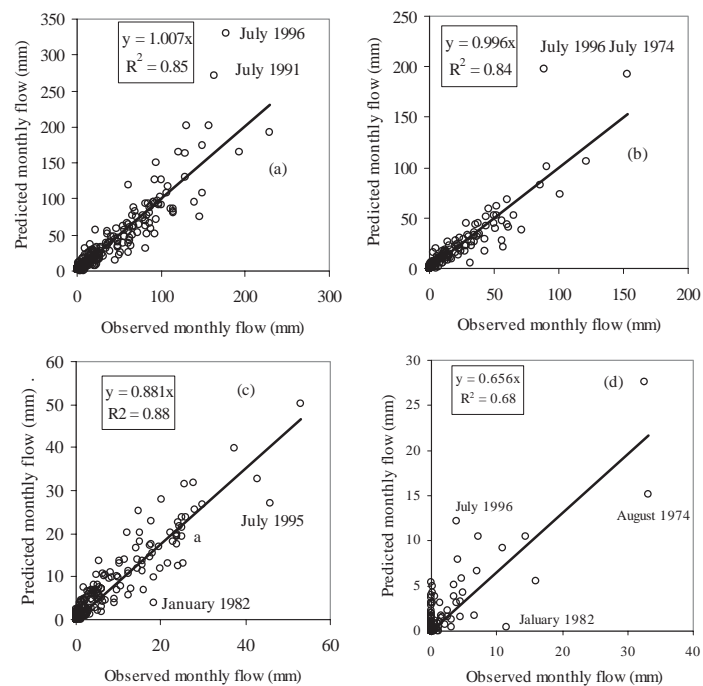


Fig. 12. Relationship between modelled and observed monthly streamflow at (a) Wights, (b) Salmon, (c) Lemon, and (d) Ernies catchments

Table 3. Water balance components (mm) of four catchments

Component	Sub-component	Wights	Salmon	Lemon	Ernies
Rainfall	Rainfall	24115	27829	17798	17798
Evapotranspiration	Interception	1660	3618	1721	2314
	Soil evaporation	2473	1236	932	102
	Transpiration	9784	19397	12218	15455
Storage change	Upper	42	27	31	0
	Lower	435	177	-1585	-549
	Groundwater	513	358	2951	249
	Stream zone	174	48	63	0
Streamflow	Surface runoff	1482	457	432	0
	Interflow	6634	2505	711	228
	Base flow	918	7	323	0
Mass balance		0	-1	1	-1

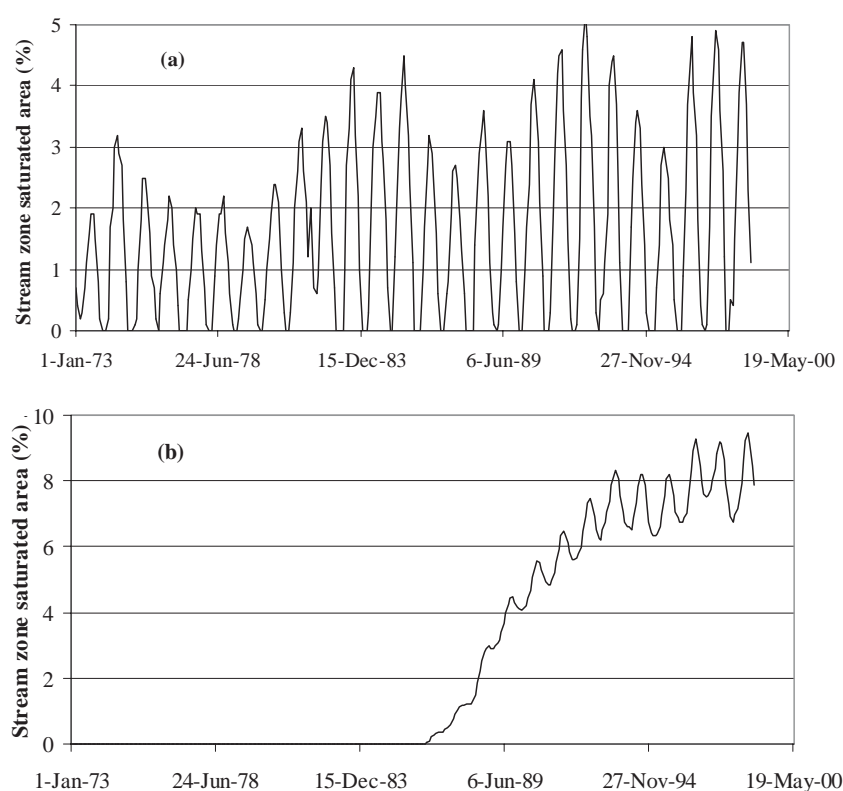


Fig. 13. Stream zone saturated areas of (a) Salmon and (b) Lemon catchments

measured. The predicted moisture content increased systematically following clearing and resulted in an increase in storage volume although the physical size of the store decreased due to rise in the groundwater level. The amount of water in the stream zone store also increased, leading to a greater saturated area. At the Lemon catchment, the

groundwater store was filled by 2951 mm, due to the largest increase in the predicted groundwater level (Fig. 7c). Similar to the Wights catchment, the lower unsaturated store decreased by 1585 mm but the predicted soil moisture content increased (Fig. 14). The transient stream zone store increased from nil to 63 mm and the saturated area increased

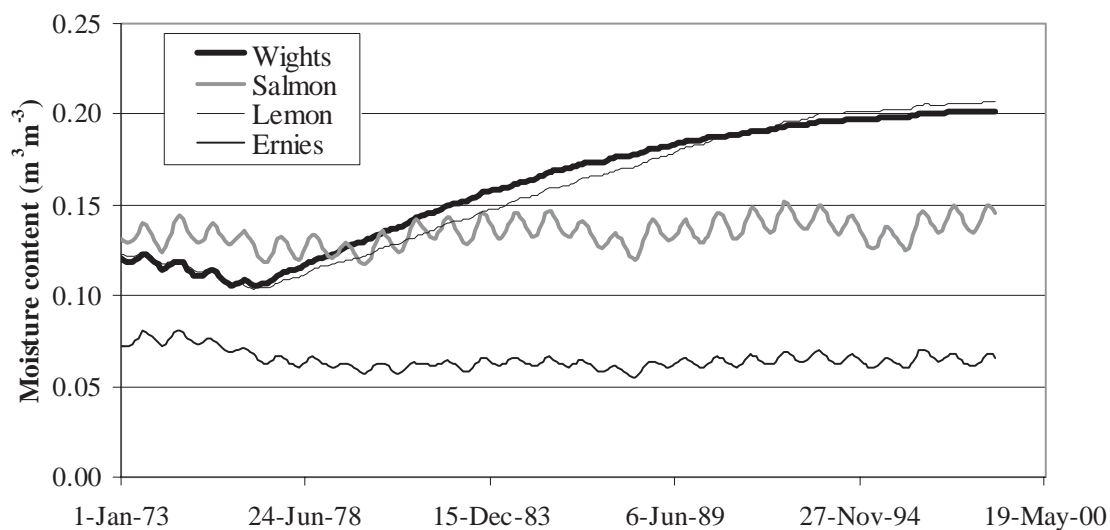


Fig. 14. Conceptual soil moisture content of lower zone unsaturated store

to 8% of the catchment area (Fig. 13b). A similar stream zone saturated area was also estimated from areal photography (Bari *et al.*, 2004).

The largest contribution to streamflow at Salmon catchment was predicted to be interflow, followed by surface runoff with base flow the least (Table 3). At Wights catchment all three streamflow components were predicted to increase following clearing, the largest relative increase being the baseflow. The surface runoff was predicted to increase more than three times due to increase in stream zone saturated area. The stream zone saturated area of 15%, as predicted by the model, is slightly less than the measured value of 20% from areal photography (Ritson *et al.*, 1995; Bari *et al.*, 2004). The interflow component was predicted to increase to a total of 6634 mm, more than 2.5 fold greater than predicted for the control (Table 3). At the Ernies catchment, streamflow is predicted to be generated only by interflow (Table 3). There are two plausible explanations: (i) there was no groundwater-induced stream zone saturated area, and (ii) the Upper Zone store never completely filled. The interflow component was predicted to more than triple at the Lemon catchment. The surface runoff and baseflow components were predicted to become active during the period of 1987–98, due to the groundwater induced saturated area (Fig. 13b).

## Discussion

The model predicted the monthly streamflow very well but over predicted for months of high-flow (high-rainfall) years. For example, the predicted streamflow in July 1996 was at least one and a half times greater than that observed at the

Wights, Salmon and Ernies catchments (Fig. 12). At the Lemon and Ernies catchments, the model also under predicted the January 1982 event. At Ernies catchment the model predicted streamflow of up to 10 mm when the stream was not flowing (Fig. 12d). Poor model prediction may be associated with sub-monthly variations in the stream zone saturated areas and thereby streamflow generation. The dynamic variation of stream zone saturated areas, due to accumulation of lateral water flow from up slope and the groundwater system, is responsible for a highly non-linear catchment response during storm events (Ruprecht and Schofield, 1989; Ruprecht and Schofield, 1991; Todini, 1995). While the prediction of the overall annual water balance from the monthly model was a slight improvement on that from the annual water balance model (Bari *et al.*, 2004), it was at the expense of model simplicity. The overall annual water balance error reduced from  $\pm 13\%$  to  $\pm 8\%$ , particularly for the Ernies and Lemon catchments. However, the annual coefficient of determination increased significantly from 0.55 to 0.83 for the Ernies catchment; this indicates an improvement in the representation of the hydrology of the catchment compared to that with the annual model.

The principal differences between the present model and LASCAM are the incorporation of a stream zone store, an explicit link of the groundwater to the stream zone and a reduction in the number of calibration parameters. In the present model the stream zone store is transient and depends upon the depth to the permanent groundwater table at the stream invert. The model has only eight parameters, most of which are physically meaningful and four remained unchanged over all catchments (Table 1).

## Summary and conclusions

A simple monthly water balance model has been developed to represent streamflow generation processes following land use changes. At first, all the collected streamflow, groundwater and other associated data were analysed from four experimental catchments (two forested control and the other two cleared for agriculture) in Western Australia. The conceptual framework of the model was then developed from data analysis and interpretation. The model was applied successfully to all four catchments and required minimal calibration.

Streamflow from both the Lemon and Wights catchments increased following clearing due to a reduction in evapotranspiration. The groundwater level declined slightly at the Salmon control catchment but it increased and enlarged the stream zone saturated area at the Wights catchment. The average annual streamflow increase, after achieving a new stability, was 300 mm at the Wights catchment. At the Lemon catchment, the groundwater level increased linearly with time and reached the stream bed in 1987, creating a 'permanent' saturated area and perennial flow with an average annual increase in streamflow of 100 mm compared to the Ernies control catchment.

In general, the impact of clearing on catchment streamflow and groundwater was captured adequately by the conceptual model. Improved predictive capability may require the incorporation of additional model complexities to capture sub-monthly responses.

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## Appendix A: symbols and variable names

$A_i$	Stream zone saturated area (mm <sup>2</sup> )		
$\Delta A_i$	Changes in stream zone saturated area (mm <sup>2</sup> )	$K_{ul}$	(mm month <sup>-1</sup> )
$A_t$	Catchment area (mm <sup>2</sup> )		Lateral hydraulic conductivity of upper store
$a_2$	Parameter related to interception (-)		(mm month <sup>-1</sup> )
$cs$	Parameter related to soil evaporation (-)	$K_{uv}$	Vertical hydraulic conductivity of upper store
$d$	Depth of top soil (mm)		(mm month <sup>-1</sup> )
$d_l$	Total depth of the soil profile (mm)	$L$	Catchment wide average stream length (mm)
$d_g^*$	Depth to groundwater level along the stream line (mm)	$L_s$	Catchment wide average hill slope length (mm)
$d_g$	Depth to groundwater level at the centre of the hill slope length (mm)	$LAI$	Leaf Area Index (-)
$\Delta d_g$	Changes in groundwater level (mm)	$LAI_{mx}$	Maximum Leaf Area Index in time (-)
$d_s$	Stream depth (mm)	$pa$	Parameter related to vertical soil conductivity (-)
$E_{su}$	Soil Evaporation from Upper Store (mm)	$pb$	Parameter related to the percolation (-)
$E_{ss}$	Soil Evaporation from stream zone store (mm)	$PET$	Daily pan evaporation (mm)
$E_{tg}$	Actual plant transpiration from Groundwater Store (mm)	$Q_i$	Shallow sub-surface flow (interflow) (mm)
$E_{tl}$	Actual plant transpiration from subsurface store (mm)	$Q_b$	Base flow (mm)
$E_{tu}$	Actual plant transpiration from Upper Store (mm)	$Q_r$	Total surface runoff (mm), ( $Q_{r1} + Q_{r2}$ )
$E_p^a$	Actual evapo-transpiration (mm)	$Q_{r1}$	Saturation excess surface runoff (mm)
$E_p$	Potential evaporation (mm)	$Q_{r2}$	Direct runoff from 'impervious area' (mm)
$I$	Percolation (mm)	$Q_t$	Total streamflow (mm)
$I_a$	Actual interception (mm)	$R$	Actual Rainfall (mm)
$ia$	Parameter related to lateral soil conductivity of A-horizon (-)	$RE$	Effective Rainfall (mm)
$K_{ll}$	Lateral hydraulic conductivity of subsurface store (mm month <sup>-1</sup> )	$RET$	Residual potential evapotranspiration (mm)
$K_{lv}$	Vertical hydraulic conductivity of the subsurface store	$RI$	Recharge to groundwater store (mm)
		$RT_g$	Root volume in the groundwater store (-)
		$RT_l$	Root volume in the subsurface store (-)
		$RT_u$	Root volume in the upper store (-)
		$RT_t$	Total root volume (-)
		$tu$	Parameter related to transpiration (-)



$W$	Water content of the upper Store (mm)	$W_{sz}$	Stream zone store water content
$W_g$	Water content of the Groundwater Store (mm)	$\Delta_{sg}$	Changes in stream zone store water content (mm)
$\Delta W_{gl}$	Change in water between subsurface and groundwater stores (mm)	$\alpha_t$	Parameter related to transpiration (-)
$W_i$	Threshold value for interflow generation (mm)	$\beta$	Catchment average groundwater slope (mm/mm)
$W_{ld}$	Water content at field capacity of the subsurface Store (mm)	$\theta_f$	Soil moisture content at field capacity (mm <sup>3</sup> /mm <sup>3</sup> )
$W_l$	Water content of the subsurface Store (mm)	$\theta_l$	Soil moisture content of subsurface unsaturated zone (mm <sup>3</sup> /mm <sup>3</sup> )
$\Delta W_l$	Changes in water content of the subsurface Store (mm)	$\theta_u$	Soil moisture content of A horizon (mm <sup>3</sup> /mm <sup>3</sup> )
$W_{mx}$	Maximum capacity of the Upper Store (mm)	$\phi_u$	Average soil porosity of A horizon (mm <sup>3</sup> /mm <sup>3</sup> )
$W_{lmx}$	Maximum capacity of the Subsurface Store (mm)	$\phi_l$	Average soil porosity of subsurface unsaturated zone (mm <sup>3</sup> /mm <sup>3</sup> )